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Title: The 2013 eruption of Chaparrastique volcano (El Salvador): Effects of magma storage, mixing, and decompression

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**Abstract:** On December 29, 2013, an isolated vulcanian-type eruption occurred at Chaparrastique volcano (El Salvador) after 12 years of inactivity. The eruption was classified as VEI 2 and produced an ash plume with a maximum height of ~9 km. Textural and compositional data from phenocrysts from the erupted products have been integrated with geochemical and isotopic information from bulk rocks to elucidate the magmatic processes responsible for the reawakening of volcanic activity. Phenocrysts consist of Fo-rich poikilitic olivines hosting high-Mg titanomagnetites, and Fo-poor olivines coexisting with low-Mg titanomagnetites. Mineral-melt equilibria suggest an origin for the distinct phenocryst populations by mixing between a high-T (~1,130-1,150 °C), basaltic magma with fO<sub>2</sub> (NNO buffer) typical of the lower crust in arc systems and a low-T (~1,060-1,080 °C), basaltic andesitic magma with fO<sub>2</sub> (NNO+1 buffer) commonly encountered in shallower, more oxidized crustal reservoirs. Thermobarometry based on Fe-Mg exchange between orthopyroxene and clinopyroxene constrains the crystallization before eruption at relative low-P (~150-250 MPa) and low-T (~1,000-1,050 °C). Mixing between two chemically distinct magmas is also evidenced by the occurrence of reverse zoned plagioclase phenocrysts with resorbed sodic cores and re-growth of sieve-textured calcic mantles. Conversely, plagioclase rims exhibit disequilibrium compositions addressed to decompression kinetics (~10-3 MPa/s) driven by rapid magma ascent to the surface (~0.03 m/s).

Major and trace element modelling excludes fractional crystallization as the primary mechanism controlling the bulk rock variability, whereas geochemical data align along a mixing trend between two end-members representative of the primitive basalt and the differentiated basaltic andesite. Trace element and isotope data indicate that the primary source of magmatism is an enriched MORB-like mantle with the contribution of fluxes of metasomatic fluids and/or melts produced by the subducted slab. The role played by slab-fluid inputs of carbonate origin and slab-melts from the hemipelagic sediments seems to be minimal. Assimilation/contamination processes of magmas by crustal rocks are also

negligible. In contrast, the geochemical signature of magmas is greatly influenced by slab-derived aqueous fluids produced prevalently by progressive dehydration of marine sediments and altered basaltic crust.

1    **The 2013 eruption of Chaparrastique volcano (El Salvador): Effects of magma**  
2    **storage, mixing, and decompression**

3  
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## 27 Abstract

28 On December 29, 2013, an isolated vulcanian-type eruption occurred at Chaparrastique  
29 volcano (El Salvador) after 12 years of inactivity. The eruption was classified as VEI 2 and  
30 produced an ash plume with a maximum height of ~9 km. Textural and compositional data from  
31 phenocrysts from the erupted products have been integrated with geochemical and isotopic  
32 information from bulk rocks to elucidate the magmatic processes responsible for the reawakening of  
33 volcanic activity.

34 Phenocrysts consist of Fo-rich poikilitic olivines hosting high-Mg titanomagnetites, and Fo-  
35 poor olivines coexisting with low-Mg titanomagnetites. Mineral-melt equilibria suggest an origin  
36 for the distinct phenocryst populations by mixing between a high- $T$  (~1,130-1,150 °C), basaltic  
37 magma with  $fO_2$  (NNO buffer) typical of the lower crust in arc systems and a low- $T$  (~1,060-  
38 1,080 °C), basaltic andesitic magma with  $fO_2$  (NNO+1 buffer) commonly encountered in shallower,  
39 more oxidized crustal reservoirs. Thermobarometry based on Fe-Mg exchange between  
40 orthopyroxene and clinopyroxene constrains the crystallization before eruption at relative low- $P$   
41 (~150-250 MPa) and low- $T$  (~1,000-1,050 °C). Mixing between two chemically distinct magmas is  
42 also evidenced by the occurrence of reverse zoned plagioclase phenocrysts with resorbed sodic  
43 cores and re-growth of sieve-textured calcic mantles. Conversely, plagioclase rims exhibit  
44 disequilibrium compositions addressed to decompression kinetics (~10<sup>-3</sup> MPa/s) driven by rapid  
45 magma ascent to the surface (~0.03 m/s).

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47 mechanism controlling the bulk rock variability, whereas geochemical data align along a mixing  
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49 andesite. Trace element and isotope data indicate that the primary source of magmatism is an  
50 enriched MORB-like mantle with the contribution of fluxes of metasomatic fluids and/or melts  
51 produced by the subducted slab. The role played by slab-fluid inputs of carbonate origin and slab-  
52 melts from the hemipelagic sediments seems to be minimal. Assimilation/contamination processes

53 of magmas by crustal rocks are also negligible. In contrast, the geochemical signature of magmas is  
54 greatly influenced by slab-derived aqueous fluids produced prevalently by progressive dehydration  
55 of marine sediments and altered basaltic crust.

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57 Keywords: Chaparrastique volcano; Central American volcanic arc; magma mixing; magma  
58 decompression;

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## 79    **1. Introduction**

80            El Salvador is one of the most seismically-active regions on Earth (Fig. 1a), being located  
81    along the Central American volcanic front that is defined by ~50 major volcanic centers, including  
82    composite or clustered volcanoes, distributed nearly continuously along 1,200 km from western  
83    Guatemala to central Panama (e.g., Carr et al., 1979; Carr, 1984; MacMillan et al., 2004). The  
84    Volcanic Cordillera in El Salvador is a line of volcanoes parallel to the coast and the offshore  
85    subduction zone. El Salvador has 23 volcanoes and 5 volcanic fields. At present, only a few  
86    volcanoes are active (outgassing and rarely ash emissions) such as the Santa Ana, San Salvador,  
87    San Miguel and Izalco volcanoes (e.g., Agostini et al., 2006 and references therein).

88            Chaparrastique volcano, also known as San Miguel, is a composite volcano in central-  
89    eastern El Salvador, ~15 km southwest of the city of San Miguel. A rural population of ~70,000  
90    people live within 10 km of the summit crater of Chaparrastique, and San Miguel is the second  
91    largest city in El Salvador. On December 29, 2013, after 12 years of inactivity, Chaparrastique  
92    volcano erupted at 10:30 local time (16.30 GMT), prompting the evacuation of more than ~5,000  
93    people living in ~3 km radius around the volcano (Martinez-Hackert et al., 2015). The eruption was  
94    a vulcanian-type explosion that lasted 2.5 hours (Fig. 1b). The explosion was classified as VEI 2  
95    and produced an ash plume with a maximum height of ~9 km (Martinez-Hackert et al., 2015). The  
96    eruption column generated heavy ash fall in nearby areas downwind, especially in the towns of  
97    Chinameca and San Jorge. A short lived pyroclastic density current generated from an early  
98    Vulcanian explosion travelled ~500 m down the flanks of the volcano and entered coffee  
99    plantations.

100            On January 2014, in response to a request of support by the government of El Salvador, the  
101    INGV (Istituto Nazionale di Geofisica e Vulcanologia, Italy) organized a task force in close  
102    collaboration with volcanologists from MARN (Ministerio de Medio Ambiente y Recursos  
103    Naturales, El Salvador). A campaign survey was conducted over a period of ten days, with the aim  
104    to install a monitoring network on the flanks of the volcano (Bonforte et al., 2015; Granieri et al.,

105 2015; Scarlato et al., 2014). At the same time, a suite of twelve eruptive products (i.e., fallout  
106 tephra) was collected in correspondence of the seismic monitoring stations of VSMG, RANC, and  
107 LACA located, respectively, at 1 km north, 2 km east, and 3 km south-east to the crater (Fig. 1c).

108 Here we present textural and mineral chemistry data from phenocrysts integrated with bulk  
109 rock geochemical and isotopic data, with the aim to elucidate magmatic processes that triggered the  
110 2013 eruption. Mineral-melt equilibria, and major and trace element modelling define the pre-  
111 eruptive crystallization conditions of the system and indicate that magma mixing and magma  
112 decompression were the most important mechanisms for triggering the eruption. The magma source  
113 has been also constrained by trace element and isotope systematics. Magmas were generated by  
114 partial melting of an enriched MORB-like mantle wedge with a dominant contribution of aqueous  
115 fluids derived from the subducted slab.

116

## 117 **2. Geological setting**

118 The Pacific coastline of much of Central America is marked by a line of active and quiescent  
119 volcanoes known as the Central American Volcanic Arc (CAVA) (e.g. Aubouin et al. 1982; Carr et  
120 al. 1990, 2004; Leeman et al. 1994; Protti et al. 1995). To the south of Central America the Cocos  
121 Plate, which underlies an area of the east Pacific, is being subducted beneath the Caribbean Plate  
122 along the Middle American trench. The volcanoes result from the upward movement of magma  
123 generated along the subduction zone between the Cocos and Caribbean tectonic plates. El Salvador,  
124 and neighbouring Central American states, lies on the Caribbean Plate. Frequent earthquakes also  
125 occur along the plate boundary. From Guatemala to northern Costa Rica, ~25 Ma old crust (formed  
126 at the East Pacific Rise) is subducted at an angle varying from 55° in Guatemala and El Salvador to  
127 65° in Nicaragua and northern Costa Rica (Protti et al. 1995; Syracuse and Abers 2006). The crustal  
128 thickness, angle of subduction, and convergence rate contribute to the thermal structure of the  
129 subduction zone, which is a major control on the generation of magmas in Central America (Davies  
130 and Stevenson, 1992). The sediment input consists of a ~200-m-thick layer of hemipelagic clay



131 overlying a ~250-m-thick layer of carbonate ooze (Aubouin et al. 1982; Plank and Langmuir 1998).  
132 The calc-alkaline compositions of volcanic rocks from CAVA record systematic along-strike  
133 variations in many geochemical tracers of the slab, including Ba/La (Carr et al., 1990), U/Th  
134 (Patino et al., 2000), B/Be (Leeman et al., 1994), and  $\delta^{18}\text{O}$  (Eiler et al., 2005). For most of these  
135 tracers, the peak in inferred slab flux occurs in the central portion of the CAVA, in Nicaragua, and  
136 falls to nearly the global minimum to the southeast, at the terminal sector of volcanoes of Costa  
137 Rica. Where the crust is thinner, high-Mg basalts are more abundant (e.g., Nicaragua). Conversely,  
138 in correspondence of a thick and old continental crust, low-Mg basaltic lavas occur and are less  
139 abundant (e.g., central and western Guatemala). Moreover, the continental crust of southern Central  
140 America is relatively young and thin, compared to the crust of other continental margins, making it  
141 less radiogenic (Carr 1984; Donnelly et al. 1990).

142 Chaparrastique volcano in eastern El Salvador is a 2,130 m-high symmetrical stratovolcano  
143 with a broad, deep crater complex, which has frequently changed morphology during historic  
144 eruptions (Escobar, 2003). Chaparrastique is considered one of the most active volcanoes in El  
145 Salvador, with 26 eruptions in the past 500 years. Seismic activity is high, while surface activity  
146 consists of pulsating gas emissions that ascend to maximum heights of ~200 m. On December 29,  
147 2013, the volcano erupted ~80 vol.% of crystal-free juvenile material (shards of fresh glass and  
148 crystal fragments) associated with ~20 vol.% of lapilli and scoria clasts. The eruption formed a new  
149 spatter cone in the central crater. In the days that followed, the  $\text{SO}_2$  emissions were estimated to  
150 reach 2,800 tons per day. The increasing tremor and gas emissions also suggested migration of hot  
151 fluids (gas, water, and magma) into the shallow plumbing system of the volcano. Before such event,  
152 the eruptive history of Chaparrastique was characterized by the occurrence of modest VEI 1–2  
153 eruptions (search for Chaparrastique at the MARN's webpage <http://www.marn.gob.sv>), mainly  
154 attributed to spattering and scoria cones at the flanks, and minor Strombolian and ash eruptions in  
155 the summit crater (Chesner et al., 2004). Since the 29 December eruption, Chaparrastique has  
156 experienced periods (from May to July 2014) of increased seismic and degassing activity, landslides

157 inside the crater, and lahars. Occasionally, small steam/phreatomagmatic explosions and minor ash  
158 falls have occurred, but no substantial explosive events have been reported. However, the level of  
159 alert remains high and explosions of small to moderate size similar to the 29 December event  
160 remain a possibility.

161

### 162 **3. Methods**

163 Textural and microchemical analyses were carried out at the HP-HT Laboratory of  
164 Experimental Volcanology and Geophysics of the Istituto Nazionale di Geofisica e Vulcanologia in  
165 Rome (Italy). Images were collected using the backscattered electron (BSE) mode of a field  
166 emission gun-scanning electron microscopy (FE-SEM) Jeol 6500F equipped with an energy-  
167 dispersive spectrometer (EDS) detector. Microchemical analyses were performed using an electron  
168 probe micro-analyzer (EPMA) Jeol-JXA8200 combined EDS-WDS (five spectrometers with twelve  
169 crystals) with 15 kV accelerating voltage and 10 nA electric current. For glasses, a slightly  
170 defocused electron beam with a size of 3  $\mu\text{m}$  was used, with a counting time of 5 s on background  
171 and 15 s on peak (Table 1S). For crystals, the beam size was 1  $\mu\text{m}$  with a counting time of 20 and  
172 10 s on peaks and background, respectively (Table 1S). The following standards have been adopted  
173 for the various chemical elements: jadeite (Si and Na), corundum (Al), forsterite (Mg), andradite  
174 (Fe), rutile (Ti), orthoclase (K), barite (Ba), apatite (P), spessartine (Mn) and chromite (Cr). Sodium  
175 and potassium were analyzed first to prevent alkali migration effects. The accuracy of the  
176 microprobe was measured through the analysis of well-characterized synthetic oxides and mineral  
177 standards. Data quality was ensured by analyzing these test materials as unknowns. Based on  
178 counting statistics, analytical uncertainties relative to their reported concentrations indicate that  
179 precision was better than 5% for all cations.

180 Bulk rock analyses for major and trace elements, as well as oxygen isotope analyses were  
181 conducted at Actlabs (Activation Laboratories Ltd.). These analyses are reported in the  
182 supplementary material (Table 2S), together with measured and certified analyses of international

reference materials. Major elements were analyzed by lithium metaborate/tetraborate fusion – ICP-OES (inductively coupled plasma optical emission spectrometry). Sample washing was performed to remove any organic, loosely adhered, and cementing material. Washing included soaking in a hot mixture of HCl and H<sub>2</sub>O<sub>2</sub> as well as cleaning in acetone using ultrasound. The analyses were performed in a batch system. Each batch contained a method reagent blank, certified reference material and 17% replicates. Samples were mixed with a flux of lithium metaborate and lithium tetraborate and fused in an induction furnace. The molten material was immediately poured into a solution of 5% nitric acid containing an internal standard, and mixed continuously until completely dissolved (about 30 minutes). The samples were run on a Thermo Jarrell-Ash ENVIRO II ICP. Calibration was performed using 7 prepared USGS and CANMET certified reference materials. FeO was determined through titration, using a cold acid digestion of ammonium metavanadate, and hydrofluoric acid in an open system. Ferrous ammonium sulphate was added after digestion and potassium dichromate was the titrating agent. This cold digestion dissolved silicates and some sulphides. Water content was determined by gravimetric method. About 0.3 g sample was thermally decomposed in a resistance furnace in a pure nitrogen environment at 110 °C (moisture, H<sub>2</sub>O<sup>-</sup>) followed by decomposition at 1,000 °C (interstitial water, H<sub>2</sub>O<sup>+</sup>), using an ELTRA CW-800, directly releasing H<sub>2</sub>O. Trace elements were measured by lithium metaborate/tetraborate fusion – ICP-MS (inductively coupled plasma mass spectrometry). Samples fused as reported above were diluted and analyzed by Perkin Elmer Sciex ELAN 6000, 6100 or 9000 ICP/MS. Three blanks and five controls (three before sample group and two after) were analyzed per group of samples. Oxygen isotopes are measured by reaction with BrF<sub>5</sub> at about 650 °C in nickel bombs. The fluorination reaction converts O in the minerals to O<sub>2</sub> gas, which is subsequently converted to CO<sub>2</sub> gas using a hot C rod. All reaction steps are quantitative. Bulk rock isotopic analyses are performed on a Finnigan MAT Delta, dual inlet, isotope ratio mass spectrometer. The data are reported in the standard delta notation as per mil deviations from V-SMOW (hereafter reported to as SMOW). External reproducibility is ±0.19‰ (1 s).

209 The isotopic compositions of Sr and Nd (Table 2S) were measured at the ETH Zürich using  
210 a Triton Plus thermal ionization mass spectrometer (TIMS). Details of analytical techniques,  
211 accuracy and precision can be found in von Quadt (1997). The isotope analysis set for Sr and Nd  
212 was performed on bulk rock powder (50–100 mg). Values for  $^{87}\text{Sr}/^{86}\text{Sr}$  were internally corrected for  
213 fractionation using a  $^{88}\text{Sr}/^{86}\text{Sr}$  value of 8.37521. Repeated measurements of the NBS 987 standard  
214 for  $^{88}\text{Sr}/^{86}\text{Sr}$  gave a mean value of  $0.710252 \pm 0.000002$  ( $2\sigma$ ).  $^{143}\text{Nd}/^{144}\text{Nd}$  values were internally  
215 corrected for mass fractionation using a  $^{146}\text{Nd}/^{144}\text{Nd}$  value of 0.7219. No correction on Nd was  
216 performed due to good segregation of Nd and Sm during column chemistry. Repeated  
217 measurements on the Nd Merck standard for  $^{146}\text{Nd}/^{144}\text{Nd}$  gave a mean value of  $0.511730 \pm$   
218  $0.000001$  ( $2\sigma$ ). The analytical uncertainties associated to the Sr and Nd isotopic compositions of our  
219 samples are in the range of 0.000006–0.000019 and 0.000002–0.000012, respectively.

220

## 221 **4. Mineralogy, petrology and geochemistry**

222 Mineral chemistry, petrology and geochemistry was performed in twelve samples of juvenile  
223 material and scoria clasts ejected during the 2013 eruption. The juvenile material consists of white-  
224 grey coarse ashes and dark-grey accretionary lapilli (i.e., CH1, CH2, CH4, CH5, CH8, CH10, CH11,  
225 and CH12 samples), and are associated to centimetre-sized scoria clasts containing phenocrysts and  
226 microlites surrounded by matrix glass (i.e., CH3, CH6, CH7, and CH9 samples). Mineral and  
227 matrix glass analyses of scoria clasts are reported in Table 1S submitted as supplementary material.  
228 Bulk rock analyses of the juvenile material and scoria clasts are reported in Table 2S submitted as  
229 supplementary material. Major and trace elements were measured for all the bulk rock samples,  
230 whereas radiogenic and stable isotopes were analyzed only for CH1, CH2, CH4, CH9, and CH12  
231 samples that represent the entire bulk rock compositional variation.

232

### 233 **4.1 Textures of scoria clasts**

234 The paragenesis of scoria clasts is dominated by phenocrysts of olivine and plagioclase (~23 vol.%),  
235 and minor pyroxene and spinel (~7 vol.%). The groundmass consists of plagioclase, pyroxene, and  
236 spinel microlites (~12 vol.%), surrounded by matrix glass (~58 vol.%). The relatively low microlite  
237 content suggests that, at the time of eruption, late crystallization was hampered by rapid quenching  
238 of magma.

239 The habit of olivine is generally euhedral with well-formed planar edges. Larger olivine  
240 phenocrysts are poikilitic and enclose sub-rounded spinels (Fig. 2a), whereas smaller olivines are  
241 accompanied by spinel phenocrysts with irregular boundaries and glass embayments resulting from  
242 the agglomeration of bleb-like grains (Fig. 2b).

243 Pyroxene crystallized as both sub-euhedral orthopyroxene and clinopyroxene.  
244 Orthopyroxenes are less frequently observed but sometimes occur as large phenocrysts partly  
245 enclosing small clinopyroxenes (Fig. 2c).

246 Plagioclase is characterized by complex zoning in which the cores, mantles and rims are  
247 separated by sharp interfaces. BSE images reveal (1) dark-grey, partly corroded cores, (2) light-grey,  
248 coarsely sieve-textured mantles, and (3) grey, euhedral to sub-euhedral rims (Fig. 2d-f). Notably,  
249 detailed BSE images of the plagioclase rims show that their final euhedral habit results from the  
250 attachment of a number of groundmass microlites growing as tabular crystals (Fig. 2g-i). These  
251 small, elongated plagioclases attach preferentially parallel preferentially parallel to the longest  
252 crystal dimension that normally corresponds to the [100] crystallographic direction contributing to  
253 the textural maturation of the rim. In contrast, plagioclase mantles are characterized by the  
254 occurrence of micrometre-sized glass channels and hollows, sometimes, in full communication with  
255 the matrix glass when the growth of the crystal rim results incomplete.

256

#### 257 4.2 Mineral and glass chemistry of scoria clasts

258 Olivine is normally zoned with forsterite (Fo) contents decreasing from core-to-rim. Larger  
259 poikilitic olivines enclosing spinels show Fo<sub>77-78</sub> cores and Fo<sub>74-76</sub> rims (Fig. 2a). The olivine core is

260 also more calcic (0.51-0.63 wt.% CaO) than the rim (0.32-0.56 wt.% CaO). Conversely, smaller  
261 olivines mutually touching with spinels in the matrix glass are weakly zoned with Fo<sub>74-75</sub> cores and  
262 Fo<sub>72-73</sub> rims (Fig. 2b).

263 Spinel is titanomagnetite with variable ulvospinel content (Usp). Smaller, sub-rounded  
264 titanomagnetites enclosed in olivine are Usp<sub>24-27</sub> (Fig. 2a), whereas single, isolated titanomagnetites  
265 are Usp<sub>21-19</sub> (Fig. 2b). Titanomagnetites in poikilitic olivines are also more primitive (~6-7 wt.%  
266 MgO) than those found in the matrix glass (~4-5 wt.% MgO).

267 Pyroxene is both orthopyroxene and clinopyroxene. Orthopyroxene exhibits weak chemical  
268 variation with a narrow enstatite (En) content of En<sub>60-63</sub> (Fig. 2c). Clinopyroxene is augite (cf.  
269 Morimoto et al., 1988) with relative low-CaO concentration (<17 wt.%) and variable diopside (Di)  
270 content ranging from Di<sub>37</sub> to Di<sub>46</sub> (Fig. 2c).

271 Among all the minerals, plagioclase phenocrysts are characterized by important  
272 compositional changes reflecting the complex textural characteristics. Plagioclases are strongly  
273 reversely zoned with resorbed sodic cores surrounded by dusty sieve-textured calcic mantles. This  
274 suggests an origin by resorption and rapid re-growth due to mixing between two compositionally  
275 distinct magmas. The anorthite (An) content sharply increases from core (An<sub>53-65</sub>) to mantle (An<sub>80-</sub>  
276 <sub>84</sub>). Conversely, at the final stage of the crystal growth, a thin-to-thick rim occurs with chemistry  
277 (An<sub>77-80</sub>) intermediates between core and mantle compositions (Fig. 2d, e, f). Importantly, the rim  
278 chemistry is almost identical to that of groundmass plagioclases, suggesting coeval crystallization  
279 (Fig. 2g, h, i).

280 The matrix glass is 52.7-54.3 wt.% SiO<sub>2</sub> and 3.5-4.6 wt.% Na<sub>2</sub>O+K<sub>2</sub>O, corresponding to the  
281 composition of a basaltic andesite (i.e., Le Bas et al., 1986). However, Mg# (100 × molecular  
282 MgO/FeO<sub>tot</sub> + MgO) values are highly variable ranging from 25 to 43. Indeed, the glass analyses  
283 are not homogeneous showing concentration gradients especially next to the well-developed rims of  
284 large plagioclase phenocrysts (see discussion below). Within such chemical variability, the

composition of the matrix glass does not differ substantially from that of micrometre-sized glass channels and glass inclusions within plagioclase phenocrysts.

#### 4.3 Bulk rock geochemistry

On anhydrous basis, the products are calc-alkaline basalts (CH1-CH2) and moderately (CH3-CH6) to more differentiated (CH7-CH12) basaltic andesites (i.e., Le Bas et al., 1986; Fig. 3a) with medium-K affinity (i.e., LeMaitre, 2002; Fig. 3b). Bulk rocks align along one single trend in which both silica and alkali explore limited compositional variations (Fig. 3a-b). From CH1 to CH12, MgO and CaO linearly decrease from ~7 to ~3 wt.% and from ~11 to ~8 wt.%, respectively, reflecting the more evolved character of the samples.

The chondrite-normalized pattern (Sun and McDonough, 1989) of REE (rare earth elements; Fig. 3c) exhibits sub-parallel trends shifting towards progressive REE enrichments from CH1 to CH12. Slight enrichments in LREE (light rare earth elements) relative to HREE (heavy rare earth elements) are also observed. Bulk rocks are characterized by weak negative Eu anomaly ( $\text{Eu}/\text{Eu}^* = 0.86\text{-}0.92$ ) and relatively low values of normalized La/Yb ratios (2.8-3.1). The primordial mantle-normalized pattern (Sun and McDonough, 1989) of trace elements (Fig. 3d) shows typical features of arc magmas, with LILE (large ion lithophile elements) to HFSE (high field strength elements) ratios higher than the primitive mantle values. This is confirmed by low Nb and Ta concentrations and correspondingly high Ba/Ta (2,900-4,200) and La/Ta (62-80) ratios. The incompatible elements generally increase with increasing  $\text{SiO}_2$ . Moreover, trace element patterns show positive spikes of Ba, U, K and Sr, and negative Nb, Ta, Ti, Hf and Zr anomalies. The compatible TE (transition elements) such as Ni (11-20 ppm) and Cr (20-58) are low and their concentration decreases from CH1 to CH12, in accordance with decreasing MgO.

Radiogenic isotope compositions of bulk rocks are constrained in narrow intervals of 0.70377-0.70387 and 0.51291-0.51299 for  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios, respectively. According to isotope data from other volcanoes in El Salvador, the lack of sensible isotopic variations with the

311 degree of magma evolution indicates that crustal contamination processes are likely to be minimal  
312 (Agostini et al., 2006).

313 Oxygen-isotope ratios range between 5.4 and 5.6 ‰, showing values comparable to most  
314 mantle peridotites ( $\sim 5.5 \pm 0.2$ ‰; e.g., Matthey et al., 1994). However, it is important to note that the  
315 oceanic lithosphere is highly heterogeneous in  $\delta^{18}\text{O}$  and may contribute with slab-derived  
316 components with oxygen-isotope ratios (0-6‰) equal to or lower than typical upper mantle  
317 peridotites, i.e., hydrothermally altered lower oceanic crust and ultramafic rocks such as  
318 serpentinites (Cocker et al., 1982; Staudigel et al., 1995).

319

## 320 **5. Discussion**

### 321 **5.1 Magma crystallization conditions**

322 In order to decipher the crystallization conditions of the erupted products, the olivine-based  
323 thermometer of Putirka (2005a) with a relatively low uncertainty of  $\pm 27$  °C has been used. Olivine-  
324 melt equilibria are particularly useful for temperature estimations because the Fe-Mg exchange  
325 reaction is nearly constant over a wide range of temperatures, bulk compositions and  $f\text{O}_2$  conditions  
326 (i.e.,  $^{\text{ol-melt}}\text{Kd}_{\text{Fe-Mg}} = 0.30 \pm 0.03$ ), and because the olivine Fo content is highly sensitive to the  
327 thermal path of magma (e.g., Roeder and Emslie, 1970). Assuming that olivine is the liquidus phase,  
328 its early-formed core is expected to be in equilibrium with the original magma represented by the  
329 bulk rock analysis (cf. Putirka 2008). Through this approach, it is found that Fo<sub>78-77</sub> olivines are in  
330 equilibrium with the primitive basalts (MgO > 6 wt.%), yielding onset temperatures of 1,135-  
331 1,146 °C (Fig. 4a). Conversely, Fo<sub>74-75</sub> olivines are in equilibrium with the more evolved basaltic  
332 andesites (MgO < 4 wt.%), recording onset temperatures of 1,067-1,084 °C (Fig. 4a). The  
333 equilibrium condition is attained when the total Fe of the melt is partitioned between  $\text{Fe}^{2+}$  and  $\text{Fe}^{3+}$   
334 through the formulation of Kress and Carmichael (1991), over buffering conditions relative to NNO  
335 (nickel-nickel oxide) and NNO+1, respectively for Fo<sub>78-77</sub> and Fo<sub>74-75</sub> olivines. These different redox  
336 states have been predicted by the spinel-melt oxygen barometer of Ariskin and Nikolaev (1996),



337 using high-Mg titanomagnetites hosted in Fo<sub>78-77</sub> poikilitic olivines (Fig. 2a) and low-Mg  
338 titanomagnetites coexisting with Fo<sub>74-75</sub> olivines (Fig. 2b). The distinct textural and chemical  
339 characteristics for olivine and titanomagnetite suggests mixing between a high-*T*, mafic melt with  
340 *f*O<sub>2</sub> typical of the lower crust of volcanic arcs (Brounce et al., 2014) and a low-*T*, differentiated melt  
341 with *f*O<sub>2</sub> commonly encountered in shallower, slightly more oxidized crustal reservoirs (see below).

342 With respect to plagioclase crystallization, the equilibrium model of Namur et al. (2012)  
343 indicates that the more sodic (An<sub>53-65</sub>) core and the more calcic (An<sub>80-84</sub>) mantle approach to  
344 equilibrium with more evolved (CaO ≤ 8 wt.%) and primitive (CaO ≥ 9 wt.%) bulk rocks,  
345 respectively (Fig. 4b). Conversely, the equilibrium condition is never found between the  
346 intermediate (An<sub>77-80</sub>) plagioclase rim and the overall bulk rock compositions. The plagioclase rim  
347 is also in disequilibrium with the surrounding matrix glass (Fig. 4b) and data deviate from the one-  
348 to-one straight line, responding to moderate An enrichments in the plagioclase rim ( $\Delta\text{An} = 16\text{-}27\%$ ).  
349 The thermometer of Putirka (2005b) with uncertainties of  $\pm 36$  °C, yields crystallization  
350 temperatures of 1,051-1,063 °C, 1,119-1,129 °C, and 1,098-1,117 °C for plagioclase core, mantle  
351 and rim, respectively. These predictions account for minimum H<sub>2</sub>O-melt concentration of 2 wt.%  
352 derived by the difference to 100% of the total of electron microprobe analyses of matrix glasses (cf.  
353 Devine et al., 1995). It has been also tested that H<sub>2</sub>O contents up to 4 wt.% changes the temperature  
354 estimates by  $\pm 32$  °C that, however, remains within the relatively high uncertainty of the  
355 thermometer. The occurrence of low-*T*, partly resorbed cores surrounded by high-*T*, sieve-textured  
356 mantles ( $\Delta T \sim 70$  °C) confirms mixing between two compositionally distinct magmas. It is therefore  
357 inferred that a colder, basaltic andesitic magma reservoir was intruded by a hotter, basaltic magma  
358 rising from depth. The increase in An content for the plagioclase mantle supports the hypothesis  
359 that the new magma was compositionally more primitive and possibly hotter and/or wetter. The  
360 textural evidence of dissolution and patchy regions at the inner part of the plagioclase corroborates  
361 the mixing mechanism (e.g., Tsuchiyama and Takahashi, 1983; Tsuchiyama, 1985; Tepley et al.,  
362 2000). Therefore, more evolved Fo<sub>74-75</sub> olivines and An<sub>53-65</sub> plagioclases crystallized from a colder

363 basaltic andesitic magma that, subsequently, mixed with the hotter basaltic magma rising from  
364 depth and carrying more primitive Fo<sub>78-77</sub> olivines. Pre-existing less calcic An<sub>53-65</sub> plagioclases in  
365 contact with the recharging basaltic magma experienced an increase in temperature that resulted in  
366 resorption textures (Fig. 3d-f). While the recharging and host magmas attempted to hybridize, high-  
367 temperature gradients lead to relatively rapid crystallization of sieve-textured, more calcic An<sub>80-84</sub>  
368 mantles with the contribution of the more primitive component of the mixing system (e.g., Hibbard,  
369 1981; Shcherbakov et al., 2010). Coherently with this scenario, the weak negative europium  
370 anomaly of the bulk rocks excludes substantial plagioclase fractionation at depth (Fig. 3c).

371 By means of the two-pyroxene model of Putirka (2008), it is found that orthopyroxene and  
372 augite exhibit Fe-Mg exchange values (i.e.,  $^{cpx-opx}Kd_{Fe-Mg} = 1.09 \pm 0.14$ ; Fig. 4c, d) indicative of  
373 equilibrium crystallization from the hybrid magma from which pyroxenes formed immediately  
374 before eruption. Temperature and pressure estimates based only on mineral chemistry (i.e.,  
375 independently on melt composition) provide values of 1,012-1,053 °C (uncertainty of  $\pm 38$  °C) and  
376 158-253 MPa (uncertainty of  $\pm 200$  MPa), respectively (Fig. 4c-d). The temperature estimates are  
377 lower than those obtained for the more evolved Fo<sub>74-75</sub> olivine and An<sub>53-65</sub> plagioclase, evidencing a  
378 late stage of pyroxene crystallization from the more oxidized hybrid magma. According to Grove  
379 and Juster (1989), if  $fO_2$  of calc-alkaline magmas increases to NNO+1, the crystallization of  
380 orthopyroxene + augite occurs at the expense of pigeonite (never observed in samples from this  
381 study) along a single reaction boundary in which olivine + liquid = orthopyroxene + augite +  
382 plagioclase. In addition, the formation of a clinopyroxene with relative high-Ca contents is also  
383 limited by previous incorporation of Ca in plagioclase.

384 Considering the wide spectrum of equilibrium experiments from the literature, the calc-  
385 alkaline melts studied by Foden and Green (1992) show compositions comparable to those from  
386 this study. At  $P \geq 500$  MPa and H<sub>2</sub>O < 2 wt.%, plagioclase is the liquidus phase followed by olivine  
387 or, alternatively, pyroxene if  $P \geq 700$  MPa. However, at  $P \geq 500$  MPa and H<sub>2</sub>O  $\geq 2$  wt.%, the  
388 stability field of liquidus olivine expands significantly, so that plagioclase does not crystallize from

389 the melt. On the other hand, at  $P = 200$  MPa and  $H_2O = 2-4$  wt.%, the crystallization sequence  
390 consists of olivine, olivine + plagioclase, and olivine + plagioclase + pyroxene (Fig. 5). This latter  
391 phase assemblage closely agrees with the textural relationships and saturation temperatures of  
392 natural minerals from this study, evidencing primary nucleation and growth of olivine + plagioclase,  
393 and subordinated formation of pyroxene. The lack of amphibole in the eruptive products is also in  
394 accordance with a low- $P$  crystallization path in which the amphibole saturation surface is  
395 encountered only at  $T < 950$  °C (Fig. 5).

396 Generally, El Salvador arc lavas crystallize at  $P \leq 500$  MPa, NNO-NNO+1 buffer, and 3–4  
397 wt.%  $H_2O$  (Agostini et al., 2006). The shallow residing conditions of magmas are corroborated by  
398 the low La/Yb ratios of bulk rocks, excluding development of large magma chambers in a thick  
399 crust with HREE-bearing minerals. Moreover, melt- $H_2O$  contents of 2-4 wt.% are frequently  
400 documented for arc magmas in Central America (e.g., Sisson and Layne, 1993; Roggensack et al.,  
401 1997; Roggensack, 2001), accounting for the remarkable fluid contribution to the mantle wedge  
402 from the subducting slab (e.g., Patino et al., 2000). Wade et al. (2006) observed that the correlation  
403 between melt- $H_2O$  contents and Ba/La ratios may be significant. On this basis, Sadofsky et al.  
404 (2007) proposed a regression fit in which the ratio of Ba/La can be used as a proxy for  $H_2O$  in  
405 magma. Through this equation it is determined that Ba/La ratios of bulk rocks from this study  
406 correspond to original melt- $H_2O$  contents of 3.3-3.7 wt.%, coherently with previous estimates for El  
407 Salvador arc lavas by Agostini et al., (2006), and the low- $P$  crystallization sequence experimentally-  
408 derived by Foden and Green (1992). Notably, LOI (Loss-On-Ignition) from bulk rock analyses  
409 (Tables 2S) ranges from 1.1-2.1 wt.% (considering total iron as FeO) suggesting overall outgassing  
410 of magma during eruption to the surface and pressure drops down to atmospheric condition (Aloisi  
411 et al., 2006; Toramaru et al., 2008; Rutherford, 2008; Gonnermann and Manga, 2013).

412

413 5.2 Magma mixing vs. magma fractional crystallization

414 Magma mixing is recognized as a widespread process in calc-alkaline, arc-related magmas that  
415 contributes to magma chamber evolution and is a likely trigger for eruptions (Sparks et al. 1977;  
416 Eichelberger et al. 2000). Previous petrological studies on volcanoes along the CAVA have  
417 documented the role of magma mixing as the cause of substantial changes in major and trace  
418 element compositions (e.g., Streck et al., 2002; Richer et al., 2004). In particular, magma mixing  
419 has been observed for a number of volcanoes in El Salvador, generally between basaltic andesitic  
420 and andesitic magmas (e.g., San Vicente volcano), and between basaltic andesitic and rhyodacitic  
421 magmas (e.g., Ilopango volcano), leading to a variety of complex rock textures characterized by  
422 corroded/resorbed plagioclases and bimodal compositional distributions of phenocrysts (e.g., Carr  
423 et al., 2007 and references therein).

424 If magma mixing occurred at the time of Chaparrastique eruption, bulk rock compositions  
425 are expected to plot on tie-lines between CH1 and CH12 end-members, resulting to be rather  
426 different from the curved trajectories caused by fractional crystallization. To test this hypothesis,  
427 MELTS (Ghiorso and Sack, 1995) simulations have been conducted using the most primitive CH1  
428 basalt as a starting liquid (Table 3s submitted as supplementary material). Thermodynamic runs  
429 started from the superliquidus temperature of 1,170 °C and continued along a fractional  
430 crystallization path that terminated at 1,020 °C. The pressure, melt-H<sub>2</sub>O content and oxygen  
431 fugacity were set at 200 MPa, 3 wt.%, and NNO+1, respectively, as the best input parameters  
432 reproducing the natural phase assemblage. Results show that the curved MELTS trajectories do not  
433 match in full with the natural Chaparrastique compositions that, in turn, align along tie-lines  
434 between CH1 and CH12 end-members (Fig. 6a). This is also true for the behaviour of trace  
435 elements modelled by the Excel spreadsheet of Ersoy and Helvacı (2010). The original partition  
436 coefficients set in the spreadsheet were used to derive the fractional crystallization vectors for REE  
437 (Fig. 6b), HFSE (Fig. 6c), and TE (Fig. 6d). While these vectors describe an implausible magmatic  
438 evolution for the Chaparrastique products, the bulk rock variability is successfully modelled by  
439 mixing (straight) lines joining CH1 and CH12 end-members (Fig. 6b-d).

440

### 441 5.3 Magma decompression and ascent to the surface

442 Plagioclase phenocrysts from Chaparrastique eruption exhibit partly resorbed An<sub>53-65</sub> cores that are  
443 the herald of early equilibrium in a basaltic andesitic magma chamber, and sieve-textured An<sub>80-84</sub>  
444 mantles interpreted as markers of further basaltic recharge (Fig. 2d, e, f and Fig. 4b). If the mixing  
445 efficiency is low, the influx of mafic magma, possibly transferring heat and H<sub>2</sub>O, causes inverse  
446 zoning of plagioclase due to dissolution and overgrowth of highly calcic crystals (e.g., Tsuchiyama  
447 1985; Davidson and Tepley 1997). Conversely, if the system rapidly re-equilibrates to form a  
448 homogeneous hybrid melt, an overall normally zoned trend is resumed with newly-formed calcic  
449 cores grading into sodic rims (e.g., Izbekov et al. 2004; Neill et al., 2015). Apparently, this latter  
450 mechanism seems to explain the final growth of An<sub>77-80</sub> plagioclase rims and the crystallization of  
451 groundmass microlites (Fig. 3d-i). However, these late-stage plagioclases are not in equilibrium  
452 with the hybrid melt if its composition is assumed to correspond with the final matrix glass (Fig.  
453 4b). Moreover, chemical profiles carried out far enough away from the plagioclase surface reveal  
454 that the matrix glass is not homogeneous (Table 4s), due to the development of chemical gradients  
455 at the crystal-melt interface (Fig. 7a).

456 Numerous cooling and decompression studies explored the effect of magma mixing and  
457 undercooling on the crystallization kinetics of plagioclase (e.g., Kirkpatrick et al., 1976; Smith &  
458 Lofgren, 1983; Muncill & Lasaga, 1987; Sato, 1995; Hammer & Rutherford, 2002; Couch, 2003;  
459 Pupier et al., 2008; Fiege et al., 2015). According to experimentally-derived data (Mollo et al., 2011,  
460 2012a), if crystal growth rate is large compared with the rate of component diffusion in the melt, a  
461 diffusive boundary layer develops causing enrichments in elements incompatible with the  
462 plagioclase crystal lattice (Fig. 7a). Concentration-dependent partitioning produces plagioclases  
463 that respond to chemical gradients in the melt, so that crystals incorporate high Ti, Mg, and Fe  
464 proportions (Fig. 7a) and become preferentially enriched in An with respect to equilibrium  
465 crystallization (Fig. 4b). The effect of undercooling due to temperature decrease (cf. Mollo et al.,

2011) and/or H<sub>2</sub>O exsolution (cf. Brugger & Hammer, 2010) leads to the formation of disequilibrium An-rich plagioclases in response to the drastic change of the topology of the Ab-An plagioclase binary loop in which the temperature difference between liquidus and solidus is minimized (cf. Iezzi et al., 2014). It is interesting to note that disequilibrium plagioclases crystallizing from rapidly decompressed calc-alkaline melts in laboratory show An enrichments ( $\Delta\text{An} = 12\text{-}30\%$ ; data from Brugger & Hammer, 2010) very close to those ( $\Delta\text{An} = 16\text{-}27\%$ ) measured in this study. The control of undercooling is also evidenced by the attachment of groundmass microlites to form the plagioclase rim (Fig. 3g-i). This crystal growth mechanism has been documented under kinetically-controlled conditions by *in-situ* experimental observations (Schiavi et al., 2009), crystal size distribution analysis (Iezzi et al., 2011; Lanzafame et al., 2013), and electron-microprobe mapping (Iezzi et al., 2014). The attachment mechanism is possibly explained by heterogeneous nucleation events on the pre-existing crystal surface (Mollo et al., 2012b; Vetere et al., 2015) and/or by the emerging theory of aggregation by self-orientation of sub-micrometric crystals (Teng et al., 2013 and references therein).

An important consequence of the diffusion-controlled growth is that the partition coefficient measured between the advancing plagioclase surface and the diffusive boundary layer changes upon the effect of chemical gradients in the melt. The partition coefficient is expressed as:

$$K_i = C_i^{xls} / C_i^{melt} \quad (1)$$

where  $C_i^{xls}$  is the concentration of a chemical element  $i$  in the crystal and  $C_i^{melt}$  is the concentration of the same element in the melt. If the compositions of both crystal and melt result from interface-controlled growth and are supposed to be in equilibrium with respect to element exchange at the crystal-melt interface, Eqn. (1) provides the ‘true’ partition coefficient  $K_i^{true}$ . Conversely, the effect of diffusion-controlled growth is to change  $K_i^{true}$  in an ‘apparent’ partition coefficient  $K_i^{app}$  when the

491 chemistry of the advancing crystal surface reflects the chemical concentrations in the diffusive  
 492 boundary layer. During plagioclase disequilibrium growth driven by undercooling, it has been  
 493 experimentally-derived that  $K_{Fe}^{app} > K_{Fe}^{true}$  and  $K_{Mg}^{app} > K_{Mg}^{true}$  (Mollo et al., 2011). For the case of  
 494 natural plagioclase phenocrysts, the lattice strain equation of Dohmen & Blundy (2014) represents a  
 495 powerful tool to predict whether or not the partitioning of Mg is controlled by equilibrium or  
 496 disequilibrium growth conditions. The model was derived by thermodynamic principles and allows  
 497 to predict  $K_{Fe}^{true}$  for a wide range of equilibrium plagioclase and melt compositions. The partitioning  
 498 of Mg measured between sodic core (or calcic mantle) of plagioclase and CaO-poor (or CaO-rich)  
 499 bulk rocks yields values close to those predicted by the lattice strain equation of Dohmen & Blundy  
 500 (2014). It is worth noting that  $\ln K_{Mg}$  increases linearly with  $1/T$ , according to thermodynamic  
 501 principles of cation partitioning (Fig. 7b). In contrast, partition coefficients measured as the ratio  
 502 between plagioclase rim and matrix glass show values much higher than those expected for  
 503 equilibrium crystallization (Fig. 7b), as for the case of  $K_{Mg}^{app} > K_{Mg}^{true}$  due to the effect of  
 504 undercooling (Mollo et al., 2011). If the advancing plagioclase rim does not violate the condition of  
 505 crystallization into an infinite melt reservoir, the value of  $K_{Mg}^{app}$  can be modelled as a function of  
 506 melt and crystal parameters (Smith et al., 1955):

$$\begin{aligned}
 K_{Mg}^{app} = 1/2 \left\{ 1 + \operatorname{erf} \left[ \sqrt{G^{pl} x / D_{Mg}^{melt}} / 2 \right] \right\} \\
 + (2K_{Mg}^{true} - 1) \exp \left\{ \left[ -K_{Mg}^{true} (1 - K_{Mg}^{true}) (G^{pl} x / D_{Mg}^{melt}) \right] \operatorname{erfc} \left[ (2K_{Mg}^{true} - 1) / 2 \sqrt{G^{pl} x / D_{Mg}^{melt}} \right] \right\}
 \end{aligned} \tag{2}$$

509  
 510 where  $K_{Mg}^{true}$  is the equilibrium partition coefficient provided by the formulation of Dohmen &  
 511 Blundy (2014),  $G^{pl}$  is the plagioclase growth velocity that is unknown,  $D_{Mg}^{melt}$  is the diffusivity of  
 512 Mg in the melt (i.e.,  $9 \times 10^{-11}$  cm<sup>2</sup>/s derived by the model of Zhang et al., 2010), and  $x$  is the

513 distance the plagioclase interface moves into the melt (i.e., the linear dimension of the plagioclase  
514 rim). Results based on the algebraic rearrangement of Eqn. (2) show that  $K_{Mg}^{app}$  measured in this  
515 study can be ascribed to rapid plagioclase growth velocities of  $\sim 2\text{--}7 \times 10^{-8}$  cm/s (Fig. 7b; see also  
516 Table 5s for the entire numerical dataset). Decompression-induced crystallization experiments of  
517 Brugger & Hammer (2010) have demonstrated that, in calc-alkaline melts, the growth velocity of  
518 plagioclase scales with the decompression path of magma ( $\Delta P$ ). It is found that  $\sim 10^{-3}$  MPa/s  $\Delta P$   
519 corresponds to  $\sim 10^{-8}$  cm/s  $G^{pl}$ . If magma decompression started at  $P = 200$  MPa (Fig. 5) over a  
520 timescale of  $\sim 10^{-3}$  MPa/s  $\Delta P$ , the ascent velocity of magma feeding eruption was  $\sim 0.03$  m/s (i.e.,  
521  $100 \text{ MPa} \approx 3 \text{ km}$ ), in agreement with values estimated for a number of volcanic eruptions  
522 worldwide (Aloisi et al., 2006; Toramaru et al., 2008; Rutherford, 2008; Gonnermann and Manga,  
523 2013; Mollo et al., 2015).

524

#### 525 5.4 Magma genesis and slab inputs

526 While partial melting of the mantle wedge is the predominant source of magmatism in Central  
527 America, the geochemical signature of arc magmas is highly influenced by the nature of the  
528 subducted slab components (e.g., Carr et al., 2004). An enriched MORB-like mantle has been  
529 identified as the primary source for CAVA magmas with the contribution of fluxes of metasomatic  
530 fluids and/or melts produced by the crustal units of the subducting Cocos Plate, i.e. the basal altered  
531 MORB unit, the middle carbonate unit, and the upper hemipelagic unit (Patino et al., 2000). A  
532 generalized approach for identifying the transfer agents between the subducted slab and the mantle  
533 wedge is based on the knowledge that ratios of some specific incompatible elements weakly change  
534 over the range of mantle melting and/or during assimilation  $\pm$  fractional crystallization processes. In  
535 contrast, the same incompatible element ratios are greatly affected by the contribution of the  
536 subducting slab, being distinctive indicators of sediment input into magma genesis (e.g., Cameron  
537 et al., 2003).



538           Considering the different geochemical signature of the hemipelagic and carbonate sediments  
539       subducted beneath Central America, the use of U/La and Ba/Th ratios has the advantage to clearly  
540       separate the two slab components (Patino et al., 2000). These ratios correspond to soluble/insoluble  
541       element pairs with the important feature that U/La is higher in the hemipelagic sediments and Ba/Th  
542       is higher in the carbonate sediments (Fig. 8a). Using partition coefficients from literature and the  
543       modal batch melting equation, Cameron et al. (2013) modelled the role played by melts and fluids  
544       derived from the subducted slab for the partial melting of the enriched MORB source (Fig. 8a).  
545       Modelling results indicate that the majority of primitive magmas at the CAVA have relatively low  
546       U/La and Ba/Th ratios and cluster near the enriched MORB source. Ba-rich fluids from the  
547       carbonate sediments produce magmas with clear slab-fluid signatures ( $>0.1\%$ ) corresponding to  
548       high Ba/Th ratios and low U/La ratios observed in Nicaragua and addressed to an increased dip of  
549       the subducting Cocos Plate (Patino et al., 2000). In contrast, as the degree of differentiation  
550       proceeds, volcanoes in Costa Rica and El Salvador are weakly influenced by slab-fluid inputs of  
551       carbonate origin (Fig. 8a). In El Salvador, Ba/Th and U/La ratios of magmas show a low  
552       contribution of melts ( $<0.04\%$ ) from the subducted hemipelagic sediments and a relative high  
553       participation of fluids ( $<2\%$ ) from the mature oceanic crust (Fig. 8a).

554           Additional important information concerning the genesis of Central American magmatism  
555       can be retrieved by using ratios of elements that are not highly fluid mobile in arc systems, such as  
556       REE and HFSE, in order to better evaluate the role of slab components with melt-like properties  
557       (Fig. 8b). Volcanic products in Nicaragua form a nearly vertical trend at low La/Sm extending  
558       towards the highest Ba/La ratios with respect to other CAVA magmas (Fig. 8b). Taken together  
559       with the discussion of Ba/Th and U/La ratios, this finding corroborates the addition to the mantle  
560       wedge of variable amounts of Ba-rich fluids derived from subducted carbonate sediments.  
561       Conversely, volcanic products in Costa Rica are characterized by the highest La/Sm and La/Yb  
562       ratios due to remarkable LREE enrichments (Fig. 8b) but among the lowest Ba/La and Ba/Th ratios,  
563       suggesting either involvement of an enriched mantle wedge or a melt-like component derived from

564 the subducted slab (Sadofsky et al., 2007). Volcanic products in El Salvador are intermediate  
565 between the geochemical signature of magmas in Nicaragua and Costa Rica, being characterized by  
566 relative low La/Sm and moderate Ba/La ratios due to pronounced slab-fluid signals from marine  
567 sediments (Fig. 8b).

568 For discriminating between crustal and source contamination of CAVA magmas, Cameron  
569 and Walker (2006) modelled the oxygen and strontium isotopes in order to define mixing  
570 trajectories between the sediment-modified mantle and crustal end-members (Fig. 8c). For most  
571 rocks in Guatemala produced by back-arc volcanism behind the volcanic front, there is clear  
572 evidence of crustal contamination (>25%). This accounts for the increased continental crust  
573 thickness and the highly radiogenic nature of the old crust (cf. Carr et al., 1990; Walker et al., 1995).  
574 In contrast, in Central America, crustal assimilation causes weak changes in the geochemical  
575 characteristics of magmas (Fig. 8c). This is particularly true for volcanoes in El Salvador,  
576 Nicaragua, and Costa Rica, emplaced in a thin and young continental crust (cf. Eiler et al., 2005)  
577 and characterized by magmatism with source contamination signature (<5%).

578 Sr and Nd systematics presented by Cameron et al. (2002) confirm that magmas in  
579 Guatemala extend in the direction of high degrees of crust contamination (Fig. 8d). On the other  
580 hand, volcanic products in El Salvador, Nicaragua, and Costa Rica define an array with positive  
581 slope (Fig. 8d) as the result of a mix between an enriched MORB source and the subducted slab  
582 (Feigenson and Carr, 1986; Carr et al., 1990). The slab source itself represents a mix between a  
583 depleted MORB source and sediments plus altered basaltic crust (Fig. 8d). Boron isotope data by  
584 Tonarini et al. (2007) for El Salvador products acknowledge the importance of slab-derived aqueous  
585 fluids as metasomatizing agents of the mantle wedge. These aqueous fluids are mostly produced by  
586 dehydration of marine sediments and altered basaltic crust of the subducted slab, but fluxes of  
587 additional fluids result also from progressive dehydration of the serpentized mantle wedge dragged  
588 beneath the volcanic arc (Tonarini et al., 2007).

589

## 590 **6. Conclusions**

591 This study documents for the first time the mineralogical, petrological and geochemical  
592 characteristics of products erupted at during the 2013 eruption of Chaparrastique volcano. Mineral-  
593 melt equilibria and bulk rock modelling indicate that the eruption was likely triggered by the input  
594 of hotter, basaltic magma from depth into a colder, basaltic andesitic reservoir residing at shallow  
595 crustal levels. Magma mixing is supported by the textural and compositional characteristics of  
596 olivine and titanomagnetite reflecting different crystallization histories, as well as by the occurrence  
597 of inversely zoned plagioclases showing partly resorbed cores surrounded by sieve-textured mantles.  
598 Furthermore, the final growth of plagioclase was controlled by the effect of undercooling due to  
599 rapid decompression during magma ascent to the surface. Trace element and isotope data tend to  
600 exclude crustal contamination, suggesting that magmas originated by partial melting of an enriched  
601 MORB-like mantle wedge metasomatized prevalently by slab-derived aqueous fluids.

602

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875

876 **Figure captions**

877 Fig.1. Schematic map showing the Central American volcanic arc (CAVA), where volcanism  
878 occurs above the plate boundary between the subducting Cocos plate and the overriding Caribbean  
879 plate (a). The ash plume of the Chaparrastique vulcanian-type eruption captured by the Suomi NPP  
880 satellite (b). Seismic monitoring stations of VSMG, RANC, and LACA located, respectively, at 1  
881 km north, 2 km east, and 3 km south-east to the volcanic crater where twelve rock samples were  
882 collected (c).

883

884 Fig.2. Textural and compositional characteristics of the scoria clasts. Larger Fo-rich poikilitic  
885 olivines include sub-rounded Mg-rich, Usp-rich titanomagnetites (a). Smaller Fo-poor olivines are  
886 accompanied by the Mg-poor, Usp-poor titanomagnetites with incomplete textures and glass  
887 embayments resulting from the agglomeration of bleb-like grains (b). Orthopyroxenes are less  
888 frequently observed and sometimes occur as larger crystals partly enclosing smaller clinopyroxenes  
889 (c). Plagioclases are characterized by dark-grey, partly corroded An<sub>53-65</sub> cores, light-grey, coarsely  
890 sieve-textured An<sub>80-84</sub> mantles, and grey, euhedral to sub-euhedral An<sub>77-80</sub> rims (d, e, f). Plagioclase  
891 rims results from the attachment of a number of groundmass microlites growing as tabular crystals  
892 (g, h, i). Ol, olivine. Timt, titanomagnetite. Opx, orthopyroxene. Cpx, clinopyroxene. Pl,  
893 plagioclase. Fo, forsterite. Usp, ulvospinel. En, enstatite. Di, diopside. An, anorthite.

894

895 Fig.3. TAS (total alkali vs. silica; Le Bas et al., 1986) diagram showing calc-alkaline basalts (CH1-  
896 CH2) and moderately (CH3-CH6) to more differentiated (CH7-CH12) basaltic andesites (a).  
897 According to the SiO<sub>2</sub> vs. K<sub>2</sub>O diagram (LeMaitre, 2002), the Chaparrastique volcanic products  
898 have medium-K affinity (b). The chondrite-normalized pattern (Sun and McDonough, 1989) of  
899 REE (rare earth elements) exhibits sub-parallel trends shifting towards progressive REE  
900 enrichments from CH1 to CH12 (c). The primordial mantle-normalized pattern (Sun and

901 McDonough, 1989) of trace elements shows typical features of arc magmas, with LILE (large ion  
902 lithophile elements) to HFSE (high field strength elements) ratios higher than the primitive mantle  
903 values (d).

904

905 Fig.4. Predictions for the crystallization temperatures of olivine phenocrysts using the thermometer  
906 of Putirka (2005a) and assuming equilibrium with the bulk rock compositions through the Fe-Mg  
907 exchange of Roeder and Emslie (1970) (a). Test for equilibrium using the plagioclase-based model  
908 of Namur et al. (2012). Temperature estimates for plagioclase are obtained by the thermometer of  
909 Putirka (2005b) (b). According to Putirka (2008), the Fe-Mg exchange between clinopyroxene and  
910 orthopyroxene suggests equilibrium crystallization conditions. The two-pyroxenes equilibrated at  
911 relative low temperatures (c) and low pressures (d).

912

913 Fig.5. Phase equilibrium experiments of Foden and Green (1992) conducted at  $P = 200$  MPa and  
914 reproducing the crystallization sequence of olivine, olivine + plagioclase, and olivine + plagioclase  
915 + pyroxene that is observed in the natural products. The experimental temperatures of Foden and  
916 Green (1992) closely agrees with the saturation temperatures predicted by olivine-, plagioclase-,  
917 and pyroxene-based thermometers, evidencing primary nucleation and growth of olivine and  
918 plagioclase, and subordinated formation of pyroxene. The lack of amphibole in the eruptive  
919 products is in accordance with low- $P$  crystallization conditions in which the amphibole saturation  
920 surface is encountered only at  $T < 950$  °C.

921

922 Fig.6. Comparison between major oxides of bulk rocks from this study and MELTS (Ghiorso and  
923 Sack, 1995) simulations conducted using as starting liquid the most primitive CH1 basalt, and  
924 initiated from the superliquidus temperature of 1170 °C and continued along a fractional  
925 crystallization path (Table 3s submitted as supplementary material). Dotted lines indicate the  
926 appearance of the mineral phase as suggested by MELTS thermodynamics. Ol, olivine. Pl,

927 plagioclase. Cpx, clinopyroxene. Opx, orthopyroxene. Sp, spinel (a). Mixing and fractional  
928 crystallization processes modelled by the Excel spreadsheet of Ersoy and Helvacı (2010). The  
929 original partition coefficients set in the spreadsheet were used to derive the fractional crystallization  
930 vectors of REE (c), HFSE (b), and TE (d). While these vectors describe an implausible magmatic  
931 evolution for the Chaparrastique products, mixing (straight) lines joining CH1 and CH12 end-  
932 members successfully model the bulk rock variability.

933

934 Fig.7. Chemical profiles carried out far enough away from the plagioclase surface (see also Table  
935 4s) reveal showing chemical gradients at the crystal-melt interface causing enrichments of cations  
936 incompatible with plagioclase crystal lattice (a). The lattice strain equation of Dohmen & Blundy  
937 (2014) predicts that, upon the effect of equilibrium crystallization,  $\ln K_{Mg}$  increases linearly with  
938  $1/T$ . The partitioning of Mg measured between plagioclase sodic core (or calcic mantle) and CaO-  
939 poor (or CaO-rich) bulk rocks yields values that closely resemble those predicted by the  
940 thermodynamic equation. In contrast, partition coefficients measured as the ratio between  
941 plagioclase rim and matrix glass show values much higher than those expected for equilibrium  
942 crystallization, as for the case  $K_{Mg}^{app}$  (i.e., disequilibrium growth)  $>$   $K_{Mg}^{true}$  (i.e., equilibrium growth)  
943 (b). Assuming that the advancing plagioclase rim does not violate the condition of crystallization  
944 into an infinite melt reservoir, the value of  $K_{Mg}^{app}$  can be modelled as a function of melt and crystal  
945 parameters and can be ascribed to rapid plagioclase growth velocities of  $\sim 2-7 \times 10^{-8}$  cm/s (c).

946

947 Fig.8. U/La vs. Ba/Th diagram based on soluble/insoluble element pairs. Using these element ratios,  
948 Cameron et al. (2002) modelled the role played by melts and fluids (grey trajectories) derived from  
949 the subducted slab for the partial melting of the mantle wedge (a). La/Sm vs. Ba/La diagram based  
950 on ratios of elements that are not highly fluid mobile in arc systems, such as the REE and HFSE.  
951 This diagram allows to evidence the contribution of to the mantle wedge of melt-like components



952 derived from the subducted slab (b).  $^{87}\text{Sr}/^{86}\text{Sr}$  vs.  $\delta^{18}\text{O}$  diagram that serve to discriminate crustal and  
953 source contamination. Cameron and Walker (2006) modelled the mixing trajectories between the  
954 sediment-modified mantle (blue trajectory) and crustal contaminants (green trajectory) (c).  $^{87}\text{Sr}/^{86}\text{Sr}$   
955 vs.  $^{143}\text{Nd}/^{144}\text{Nd}$  diagram shows evidence of crustal contamination for rocks in Guatemala that are  
956 behind the volcanic front. However, most of the volcanic products in Central America define an  
957 array with positive slope that represents a mix between the enriched MORB source and the  
958 subducted slab (d). EM, enriched MORB source. OC, oceanic crust. CS, carbonate sediments. HS,  
959 hemipelagic sediments. DM, depleted MORB source.

Figure1

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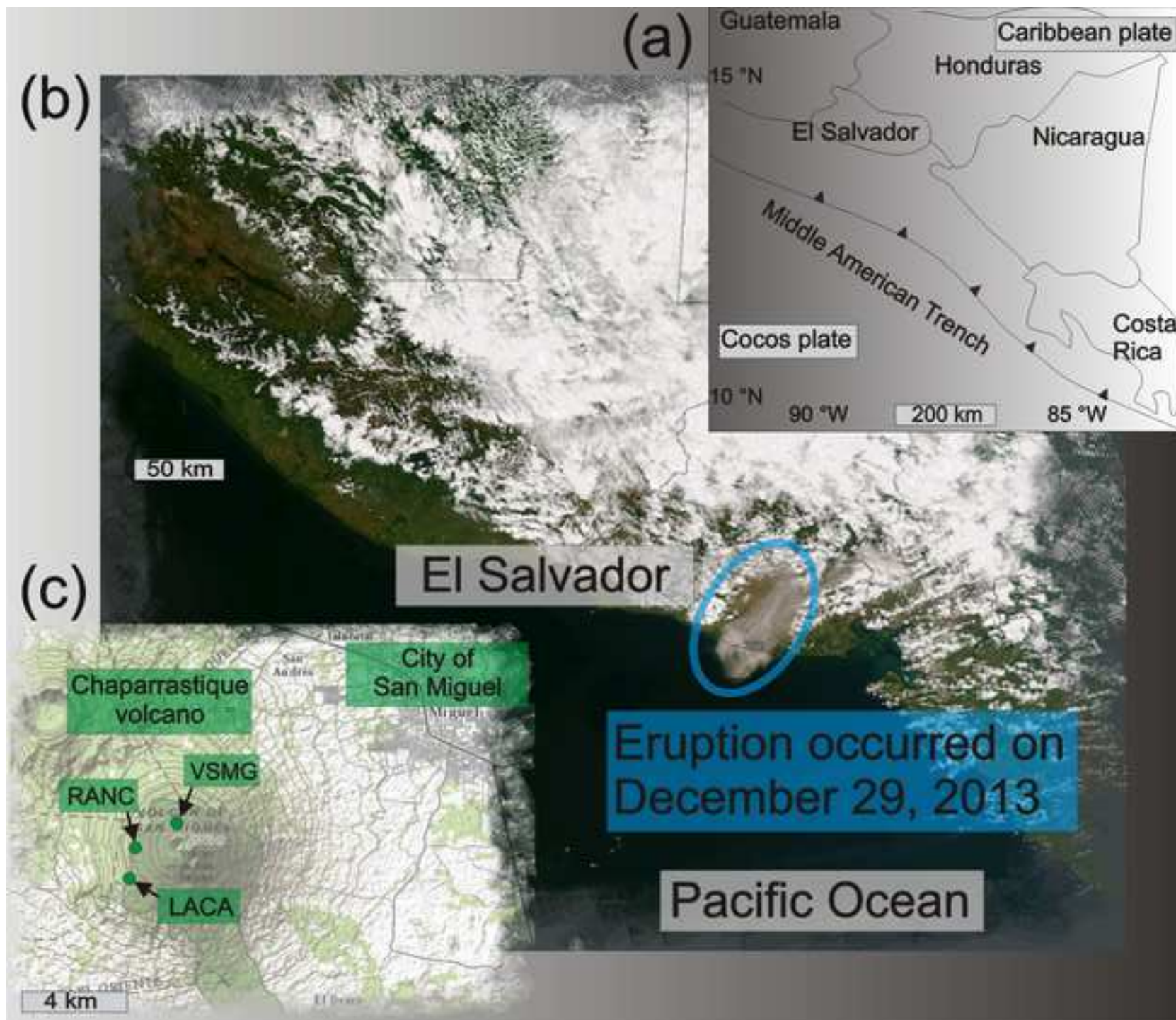




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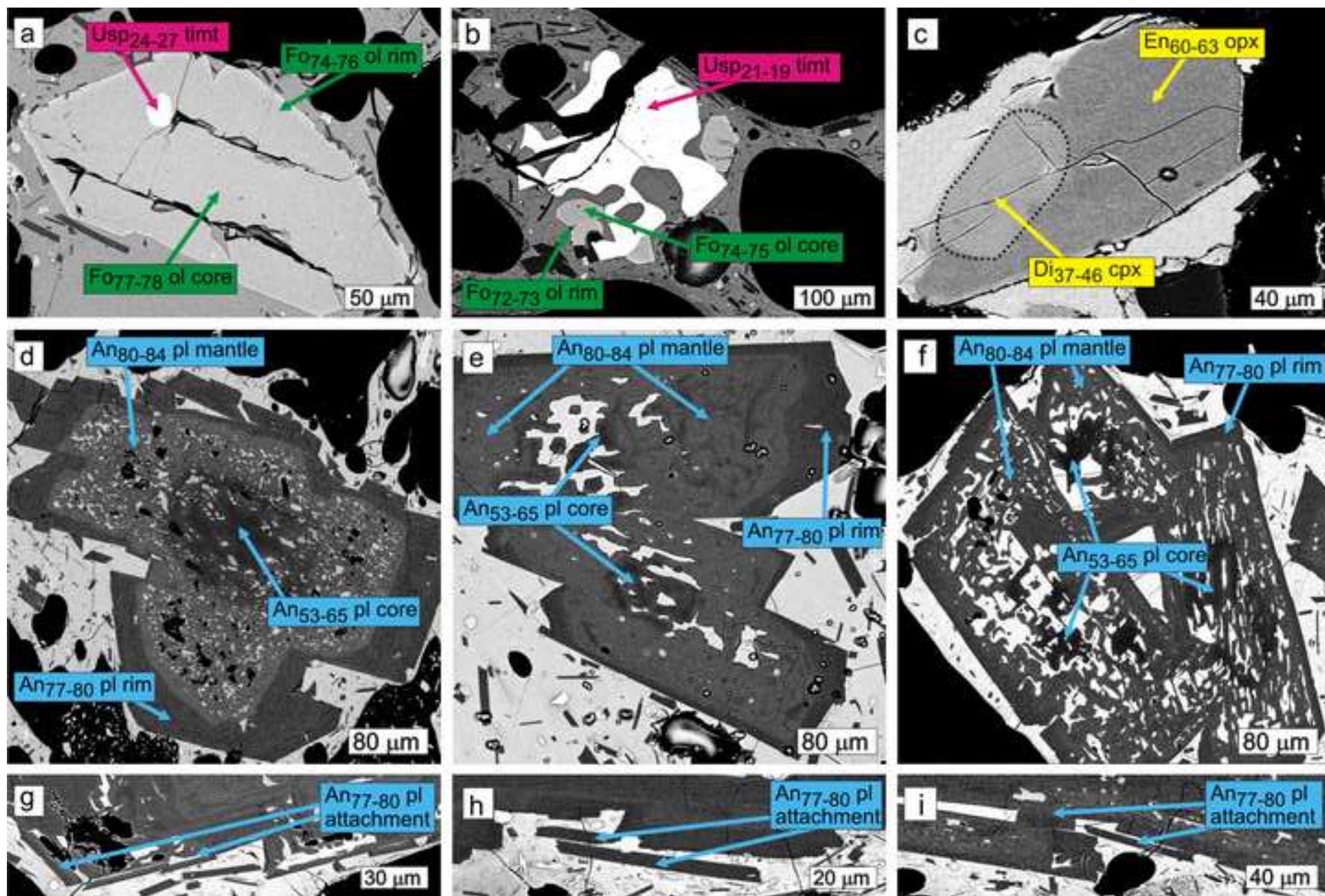


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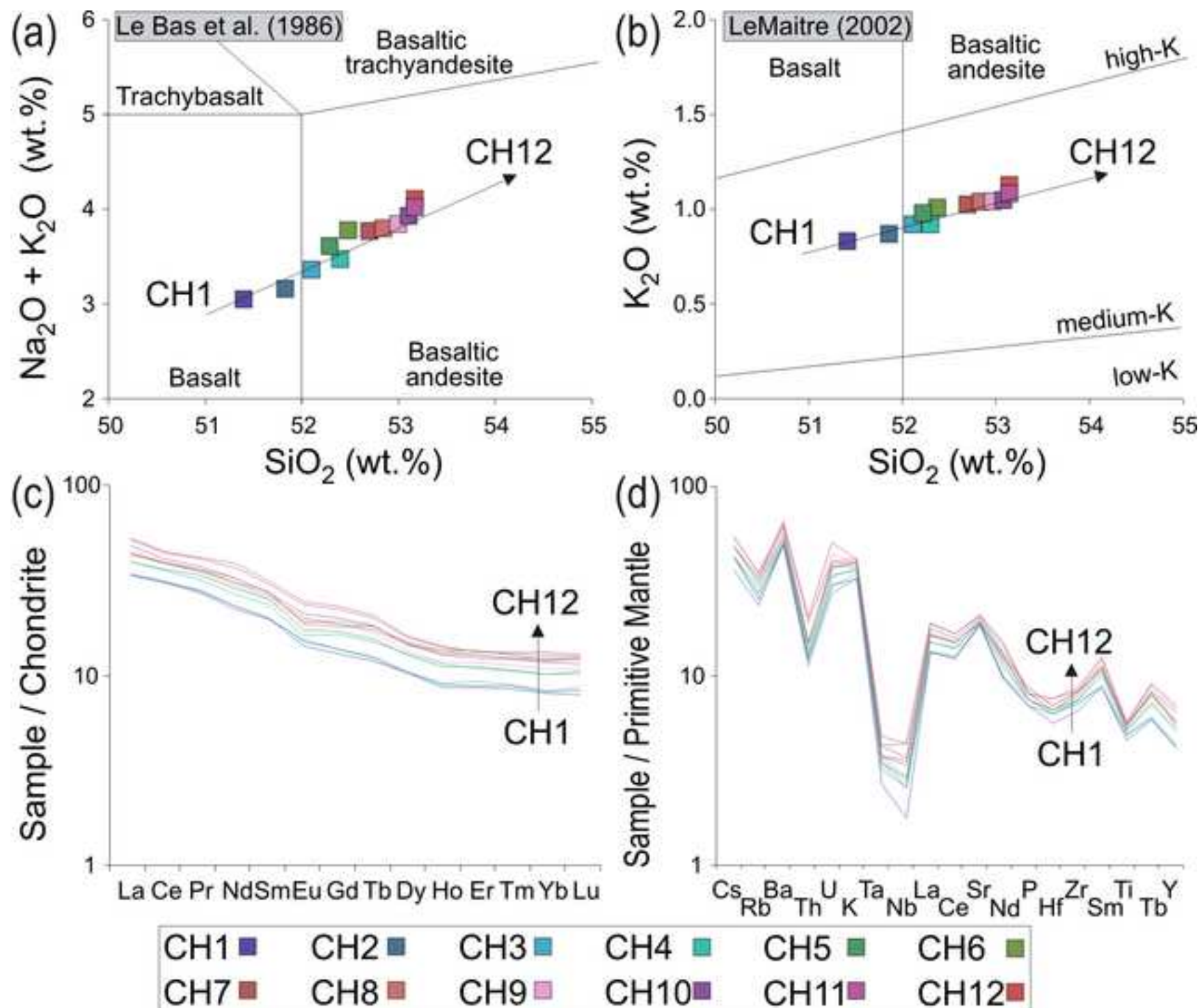




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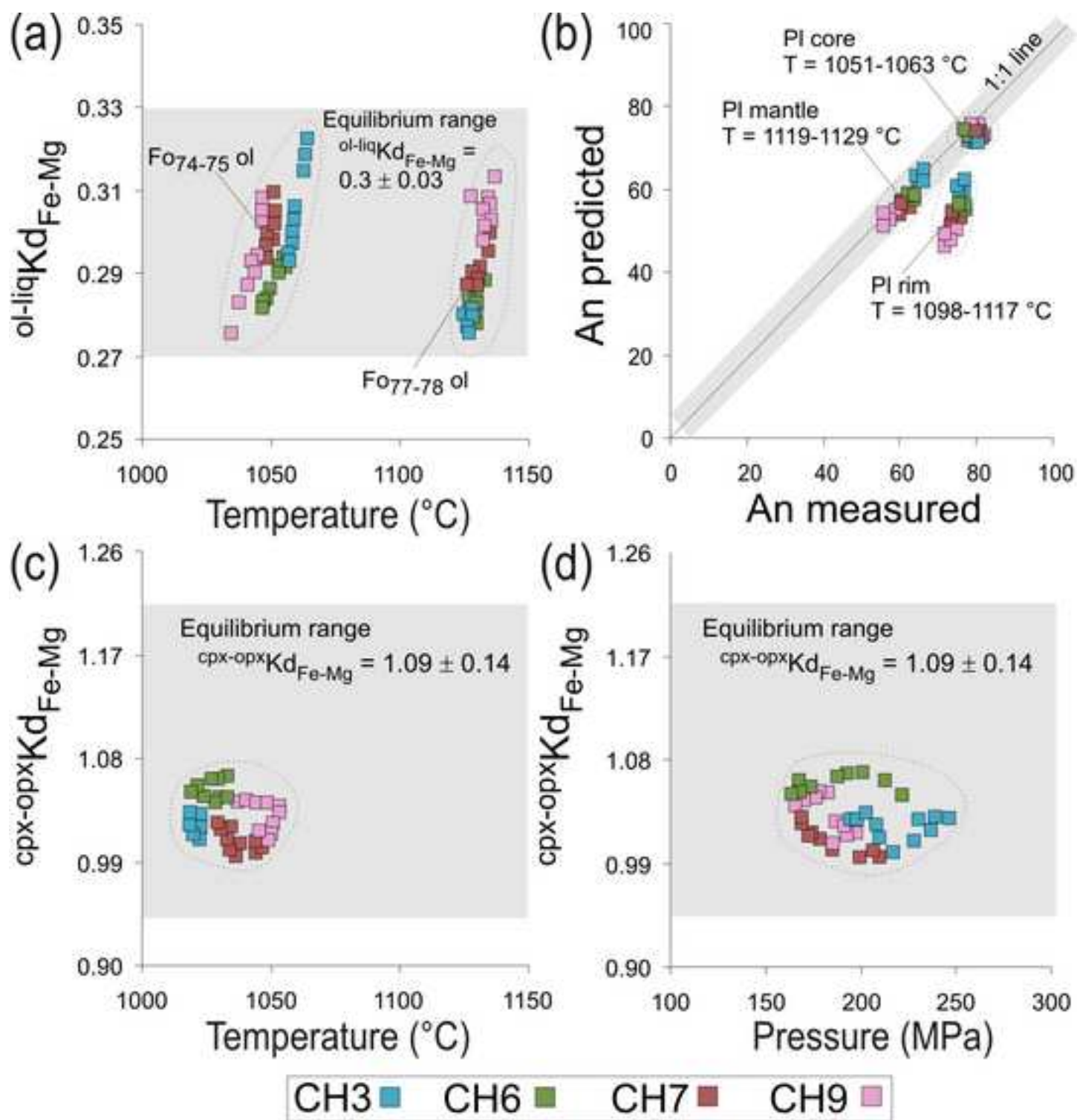


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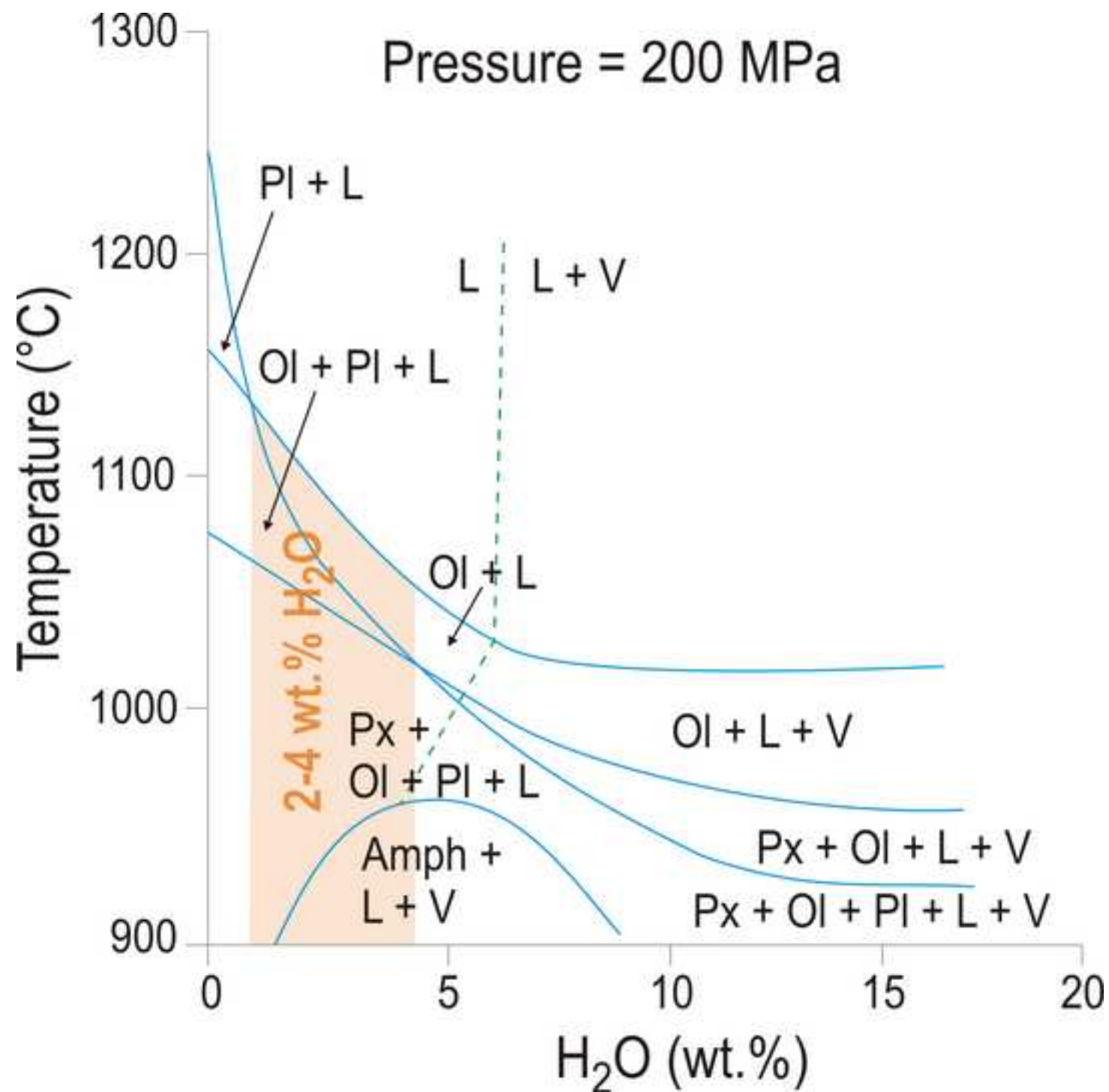


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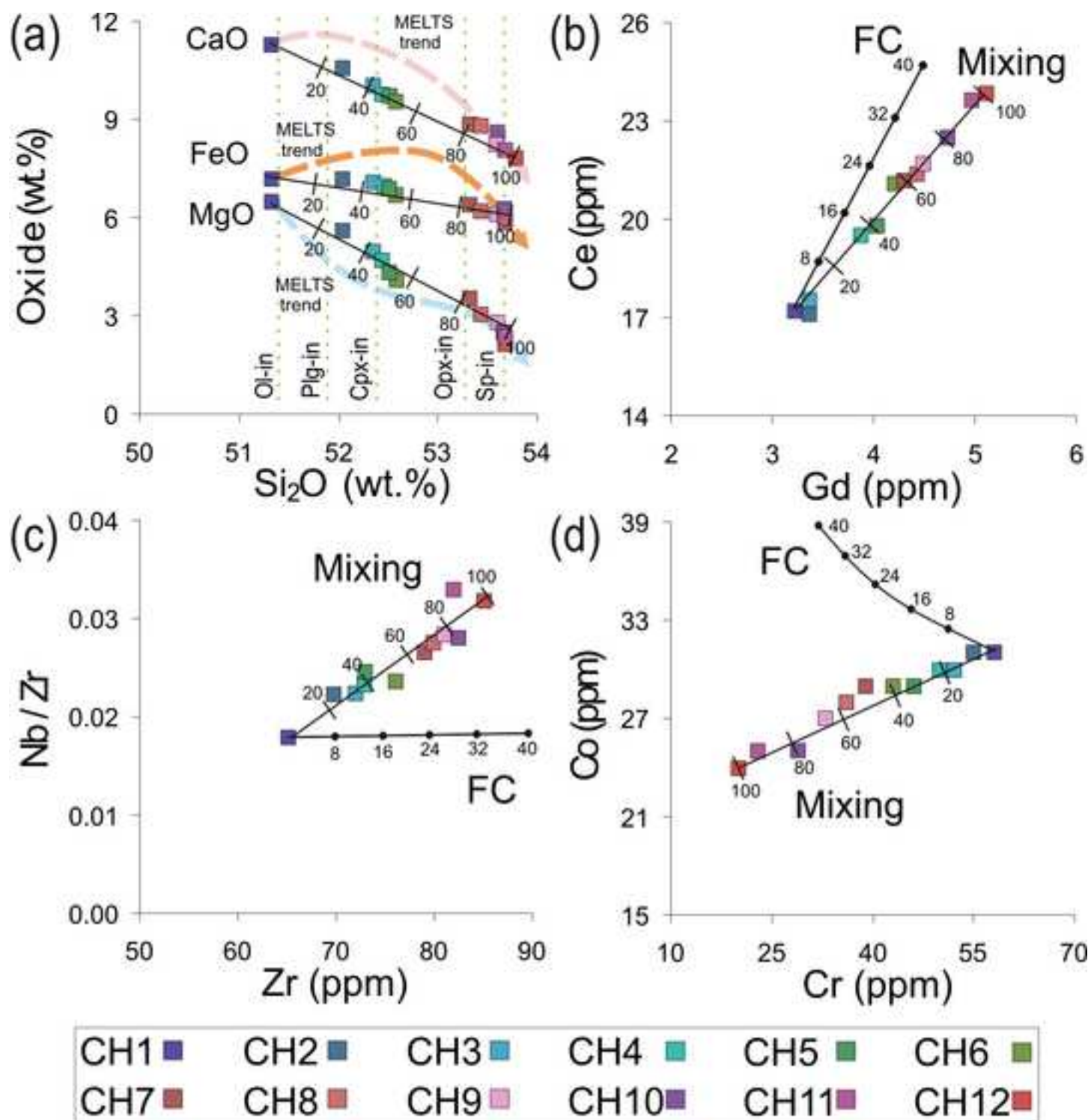


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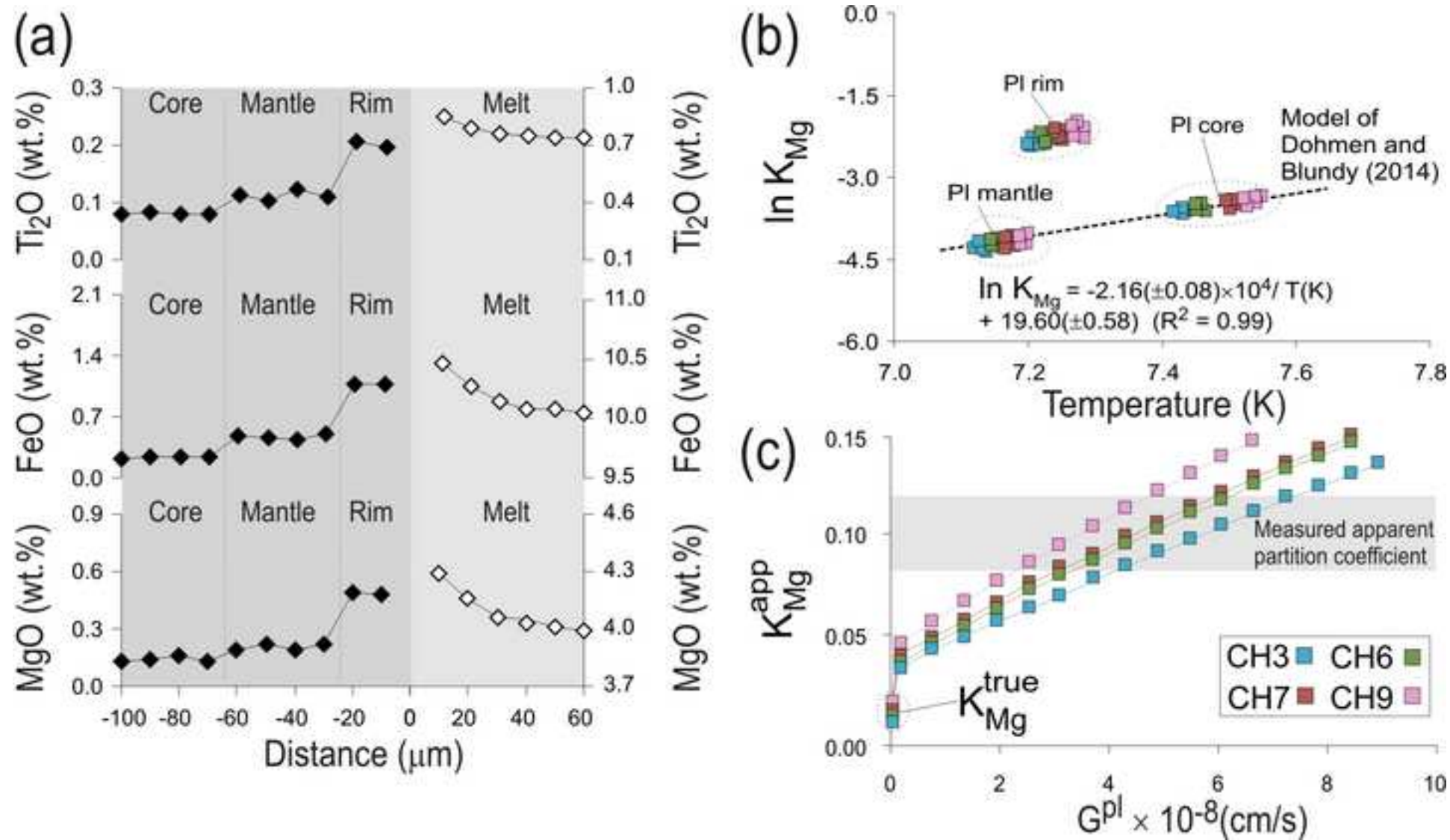
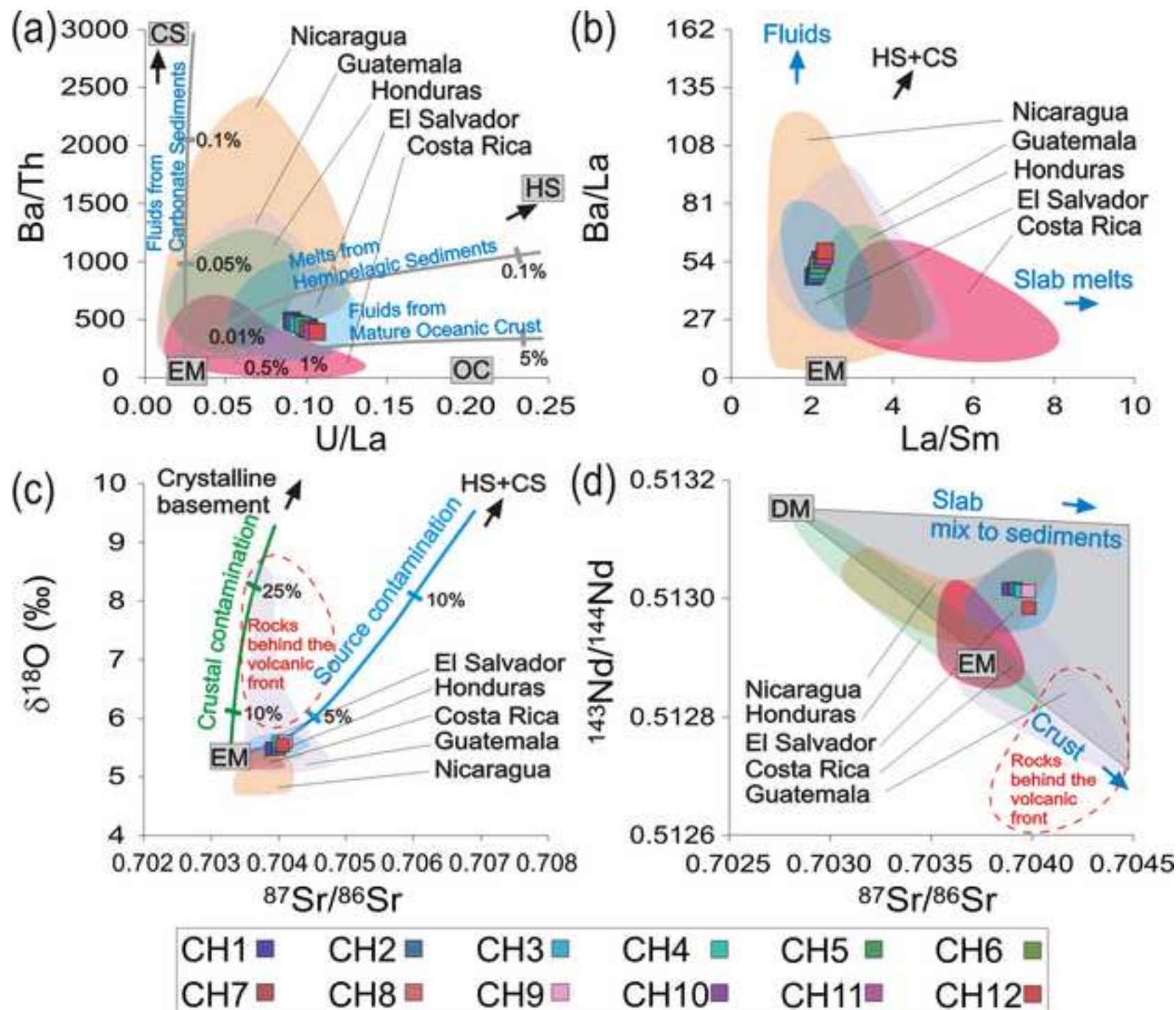
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Figure8

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Table5S

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